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UPPER AIR CIRCULATION IN LOW LATITUDES IN RELATION TO CERTAIN CLIMATOLOGICAL DISCONTINUITIES

By R. FROST, B.A.

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UPPER AIR CIRCULATION IN LOW LATITUDES IN RELATION TO CERTAIN CLIMATOLOGICAL DISCONTINUITIES

By R. Frost, B.A.

Summary.—The most striking features of the upper air circulation over Arabia are the westerly jet stream of over 100 kt. at a pressure level of about 200 mb. in latitudes 25–30°N. during the winter months and an easterly jet stream of over 70 kt. at a pressure level of 100 mb. in about latitude 15°N. during July. The westerly jet stream reaches its maximum intensity in midwinter, moves slowly northwards and decreases slightly during the spring, and then moves to a position about 40°N. and weakens rapidly during midsummer. The easterly jet stream weakens rapidly during the autumn and is non-existent during the spring and winter. As the westerly jet stream moves north over Iraq there is an abrupt change in the level of the tropopause, and it is suggested that the break-down in this tropopause barrier helps to explain the surface discontinuities of temperature and wind which occur during these months and the spectacular onset of the monsoon over India.

Introduction.—In a note in the *Meteorological Magazine* Dewar[1]* discussed the variation of the tropopause pressure and temperature at Habbaniya, and showed that during the winter months the tropopause occurred at a pressure level of approximately 220 mb. and during the summer months at a pressure level of about 80 mb. The most striking features, however, were the abrupt changes in the pressure levels which occurred in late May and to a lesser extent in October, see Fig. 2 which shows the tropopause pressure for 1950.

Climatological data for Habbaniya show that surface discontinuities of wind and temperature also occur in these months. Table I, which exhibits the monthly frequency of winds at Habbaniya, shows that during the winter months the

TABLE I—MONTHLY FREQUENCY OF WIND DIRECTIONS AT HABBANIYA
(33° 22′ N., 43° 34′ E., 144 ft.) AT 0600 G.M.T. (0900 L.T.)

Period : 10 yr., 1937–46

	Jan.	Feb.	Mar.	Apr.	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.
NNW.	24	31	41	37	49	74	73	83	73	76	37	42
N.	7	8	25	28	38	29	14	21	26	26	20	10
NNE.	3	4	4	12	12	4	1	1	6	10	7	9
NE.	4	6	8	12	13	8	1	7	6	18	21	8
ENE.	..	3	7	6	3	2	1	..	..	9	9	9
E.	12	10	12	20	12	3	1	1	1	16	15	15
ESE.	11	11	7	11	2	..	..	..	1	3	6	12
SE.	23	29	17	18	4	..	1	..	..	3	11	22
SSE.	8	10	14	10	2	2	..	..	..	..	3	5
S.	15	10	15	14	5	3	1	1	3	11	10	8
SSW.	7	5	12	7	4	1	2	..	..	2	2	5
SW.	9	10	15	6	6	2	..	1	3	9	7	8
WSW.	12	5	6	5	5	2	..	..	1	2	5	4
W.	19	20	14	11	14	6	6	3	4	2	11	16
WNW.	25	17	17	12	36	21	29	16	16	11	7	14
NW.	77	67	77	69	91	134	176	169	135	82	81	72
Calm	54	36	19	22	12	9	4	7	25	30	48	53

*The index numbers refer to the bibliography on p. 24.

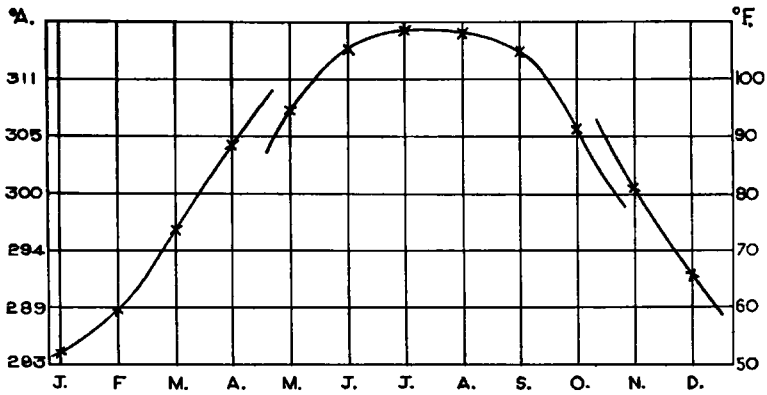


FIG. 1.—MEAN MAXIMUM SURFACE TEMPERATURE, HABBANIYA, 1950

surface winds are variable whilst during the summer months, from June to October, there is a steady NW. wind. The onset of this NW. wind, known as the shamal which is the northerly extension of the Arabian branch of the Indian monsoon, occurs in late May or early June and is frequently accompanied by duststorms; whilst not as spectacular in its arrival as the "burst of the monsoon" over India it is almost equally abrupt. Table II, which gives the mean monthly west–east and south–north components of the winds at Habbaniya at standard pressure levels, shows that during the summer months there is a northerly component of the wind at all levels from the surface to 700 mb., whereas during the winter months, with the exception of January, there is a southerly component at these levels.

Normand[2] called attention to the roughly simultaneous and similar onsets of the shamal and the monsoon, and also pointed out the discontinuous nature of the temperature in Iraq. The latter can readily be seen from Fig. 2 of Normand's paper which shows the average daily maximum temperature at Baghdad during the period 1887–1914, or from Fig. 1 of the present paper which depicts the mean maximum surface temperature at Habbaniya in 1950.

In the course of an investigation into these climatological discontinuities seasonal charts of tropopause pressure levels over the Mediterranean, Africa, Middle East and India, and seasonal meridional cross-sections for longitude 40°E. were prepared. These are reproduced as Figs. 3 and 4 and are discussed in the following sections.

It is suggested that these climatic discontinuities and also the onset of the Indian monsoon are a direct consequence of this abrupt change in the tropopause régime.

Observational material.—The main observational material used in this discussion is :—

Radio-sonde and radar-wind observations at 0300 and 1500 G.M.T. from January 1948 at

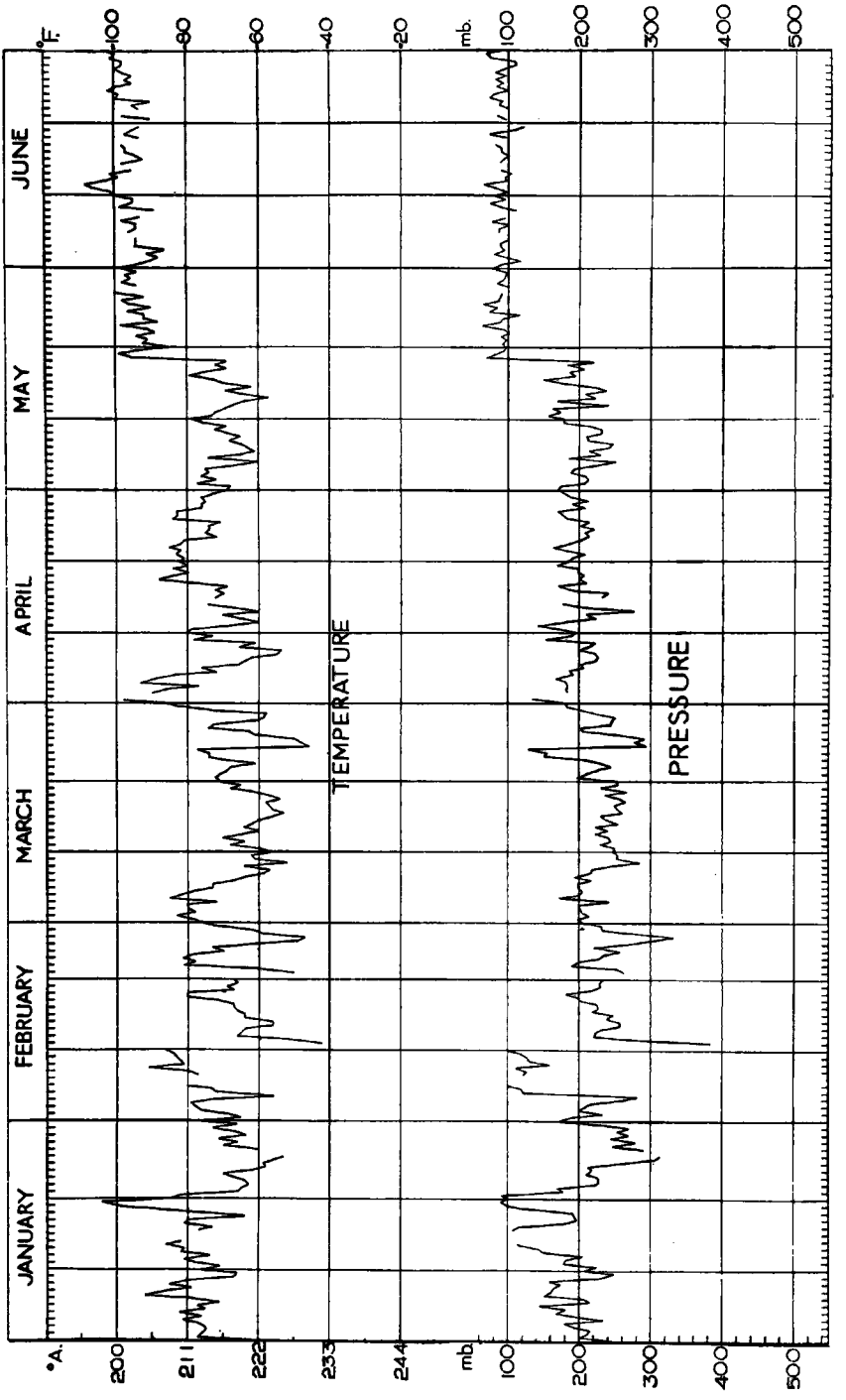
Lerwick	60°08′N. 1°11′W.
Larkhill	51°11′N. 1°48′W.
Gibraltar	36°09′N. 5°21′E.
Malta	35°50′N. 14°27′E.
Benina	32°05′N. 20°16′E.
Nicosia	35°09′N. 33°17′E.
Habbaniya	33°22′N. 43°34′E.
Bahrein	26°14′N. 50°25′E.
Aden	12°50′N. 45°01′E.

Radio-sonde and radar-wind observations at 0600 G.M.T. from January 1948 at

Nairobi	1°17′S. 36°50′E.

Radio-sonde observations for shorter periods at

Marshall Field, Liberia	6°08′N. 10°22′W.
Salisbury, Southern Rhodesia	17°50′S. 31°01′E.
Pretoria, Union of South Africa	25°46′S. 28°14′E.
Nandi, Fiji Islands	17°45′S. 177°27′E.
Canton Island	2°49′S.171°43′W.


JANUARY
FEBRUARY
MARCH
APRIL
MAY
JUNE
°F.
100
80
60
40
20
°A.
200
211
222
233
244
TEMPERATURE
mb.
100
200
300
400
500
mb.
100
200
300
400
500
PRESSURE

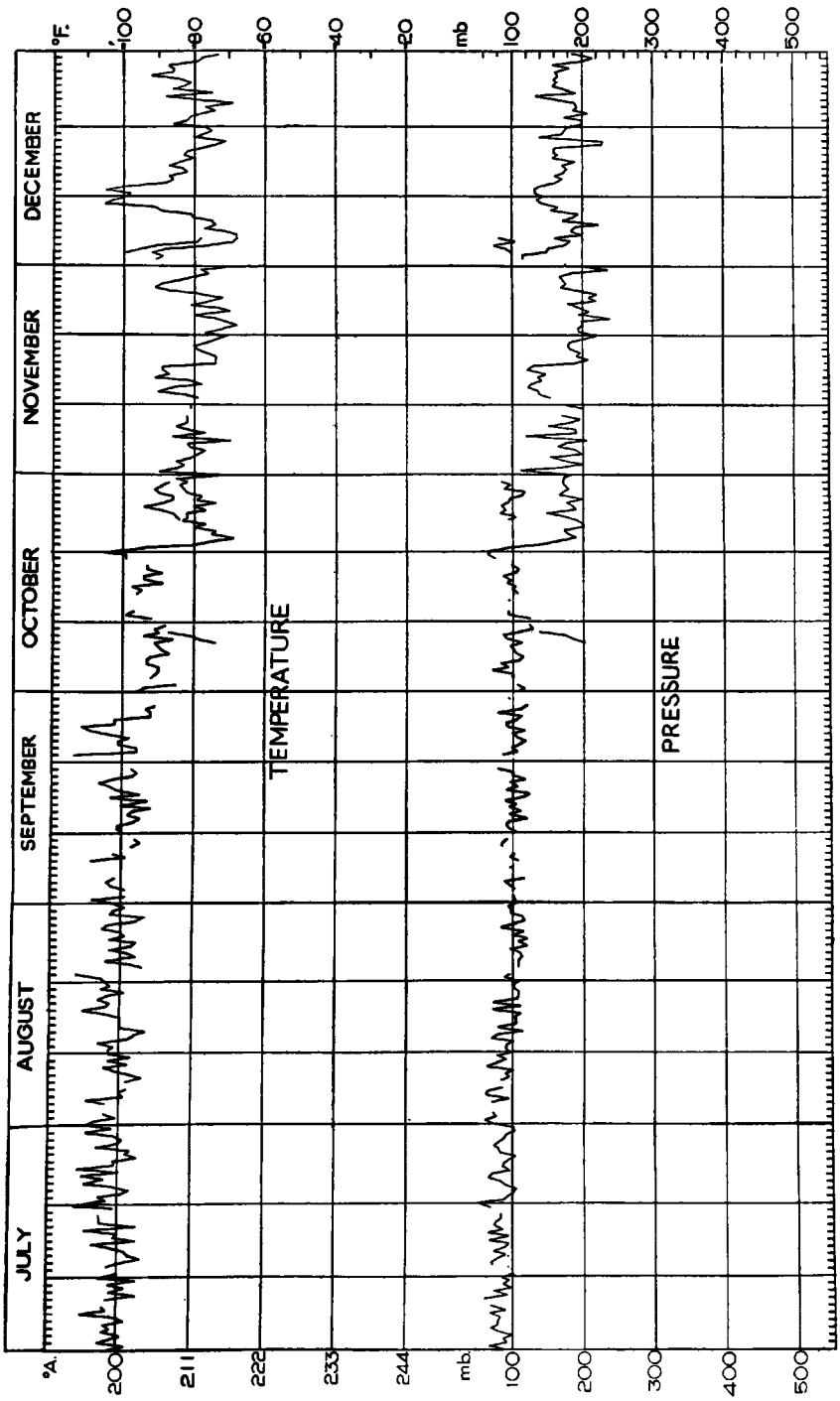


FIG. 2.—VARIATION OF THE TROPOPAUSE TEMPERATURE AND PRESSURE OVER HABBANIYA, 1950
Hours of observation 0200 and 1400 G.M.T. (0500 and 1700 Zone time)

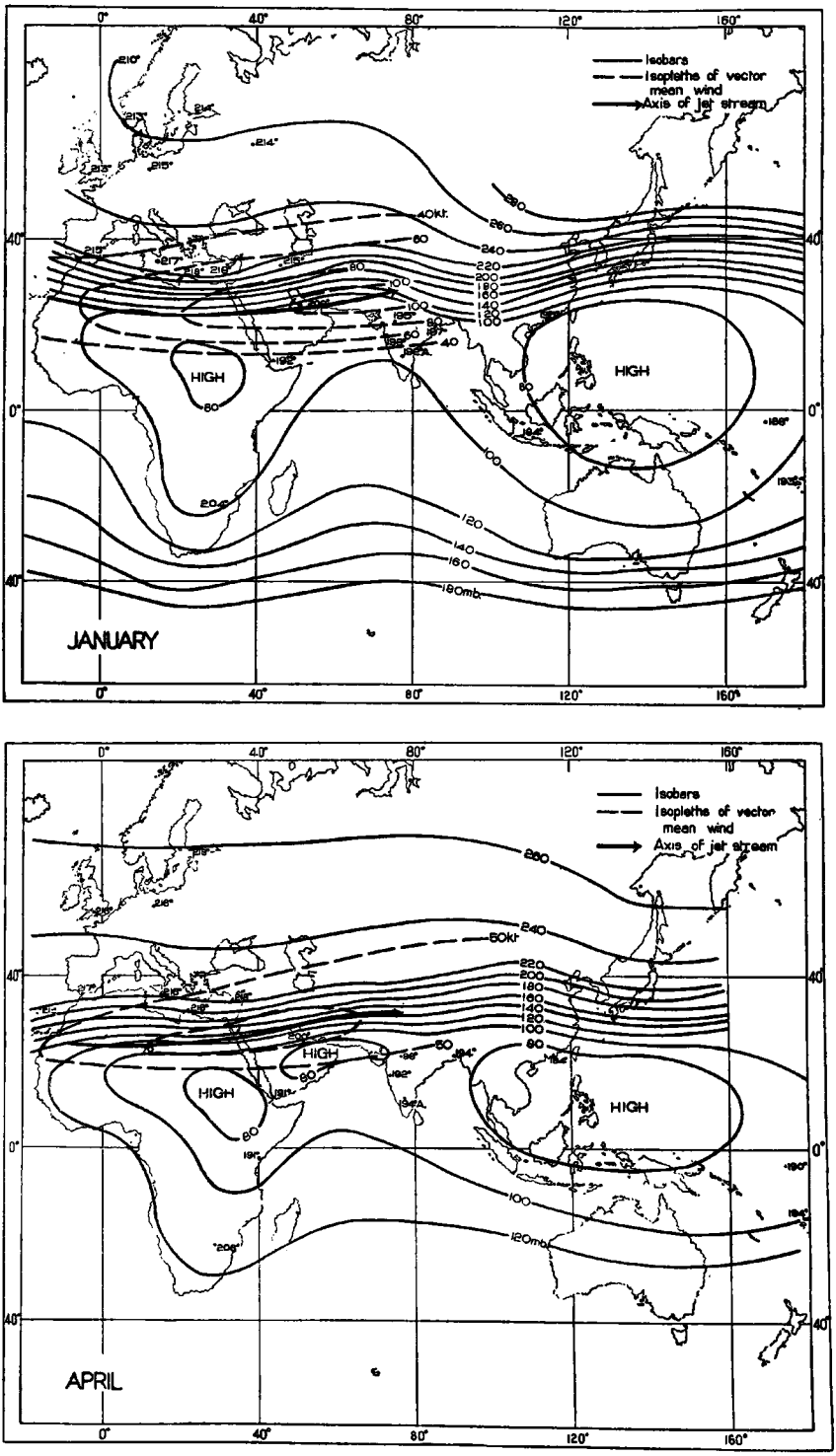
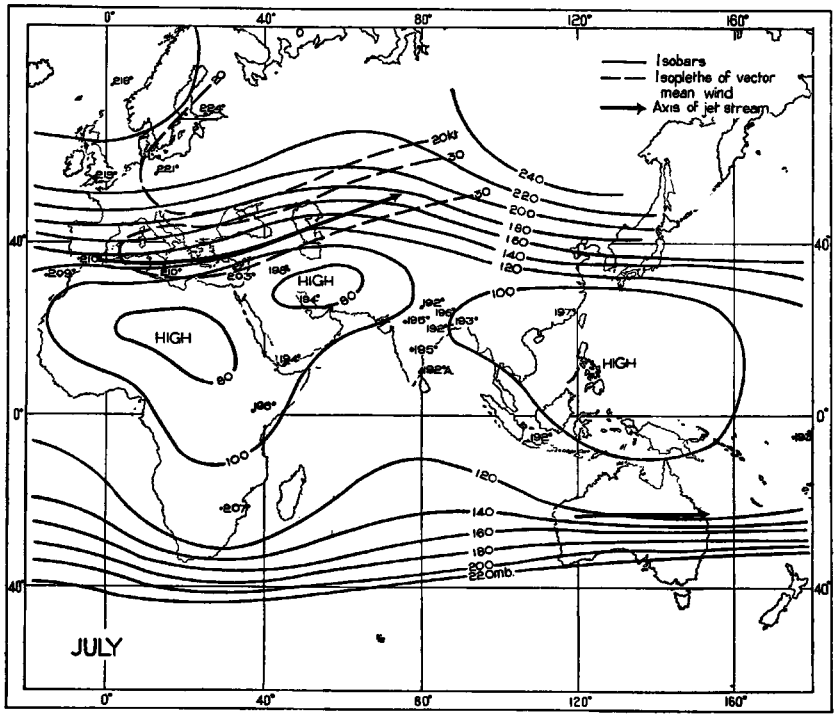


FIG. 3.—MEAN PRESSURE CHARTS OF THE TROPOPAUSE WITH ISOPLETHS OF VECTOR MEAN WIND AT 200 MB.



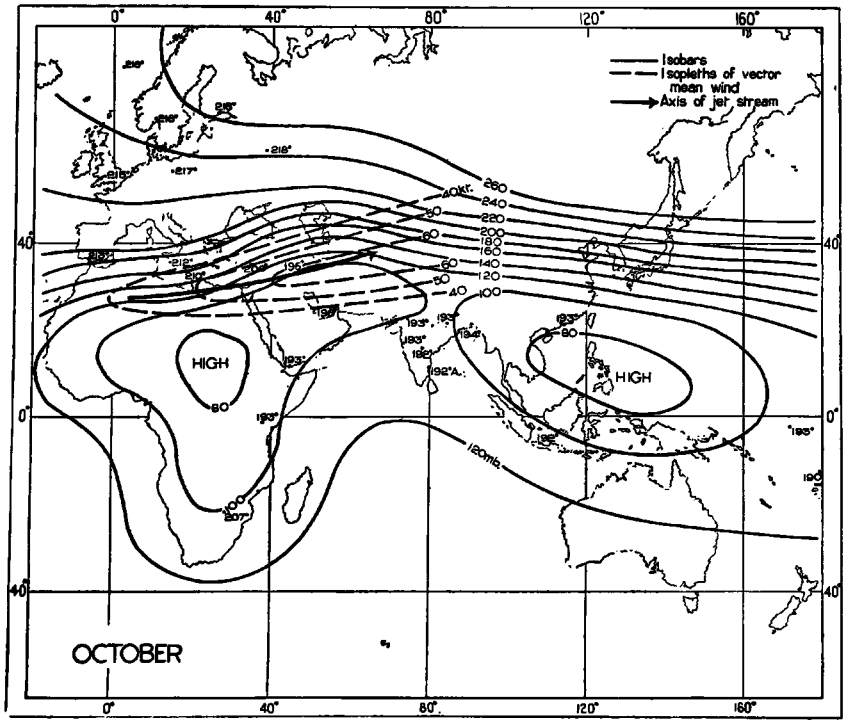


FIG. 3.—MEAN PRESSURE CHARTS OF THE TROPOPAUSE WITH ISOPLETHS OF VECTOR MEAN WIND AT 200 MB.—*continued*

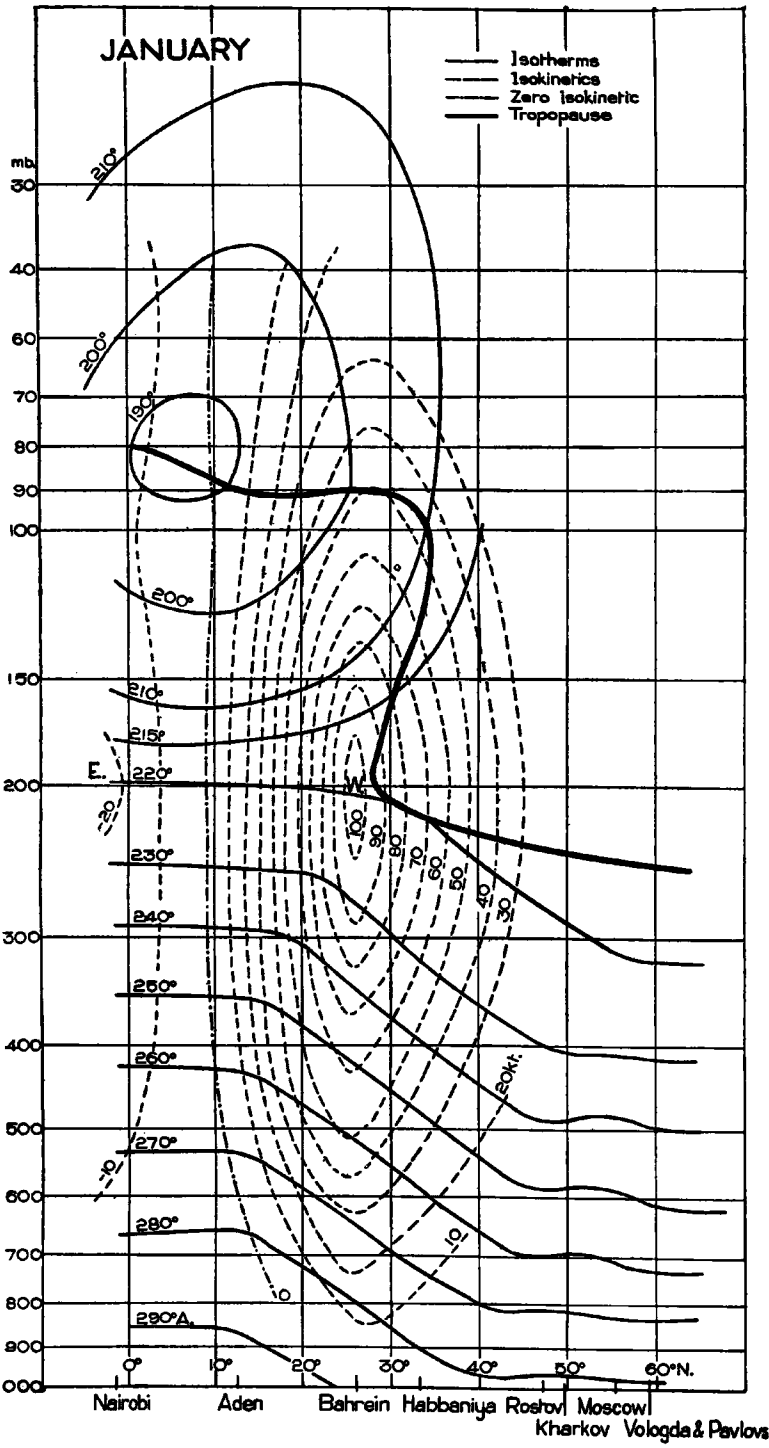


FIG. 4.—ATMOSPHERIC CROSS-SECTION AT 40°E. LONGITUDE SHOWING MEAN TEMPERATURE AND MEAN E.–W. WIND COMPONENT

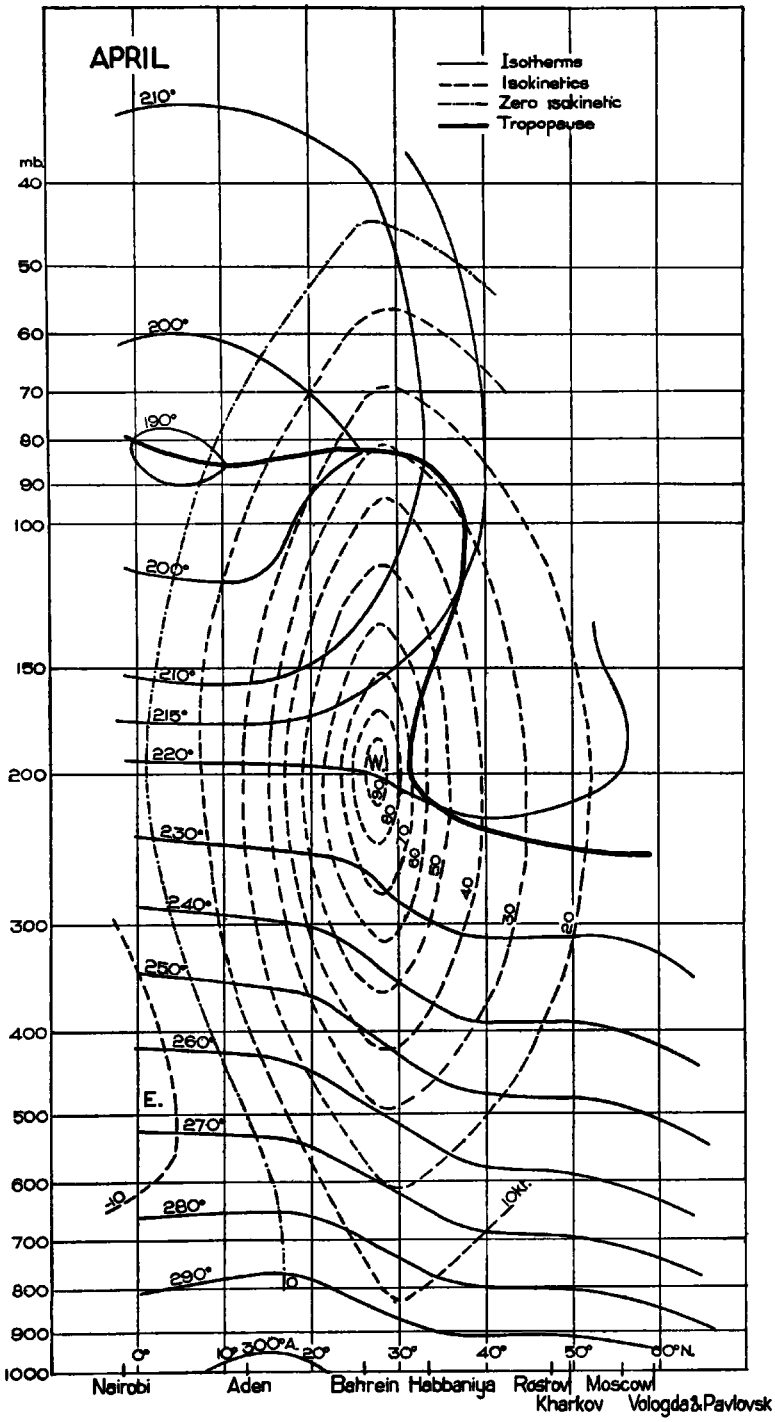


FIG. 4.—ATMOSPHERIC CROSS-SECTION AT 40°E. LONGITUDE SHOWING MEAN TEMPERATURE AND MEAN E.–W. WIND COMPONENT—*continued*

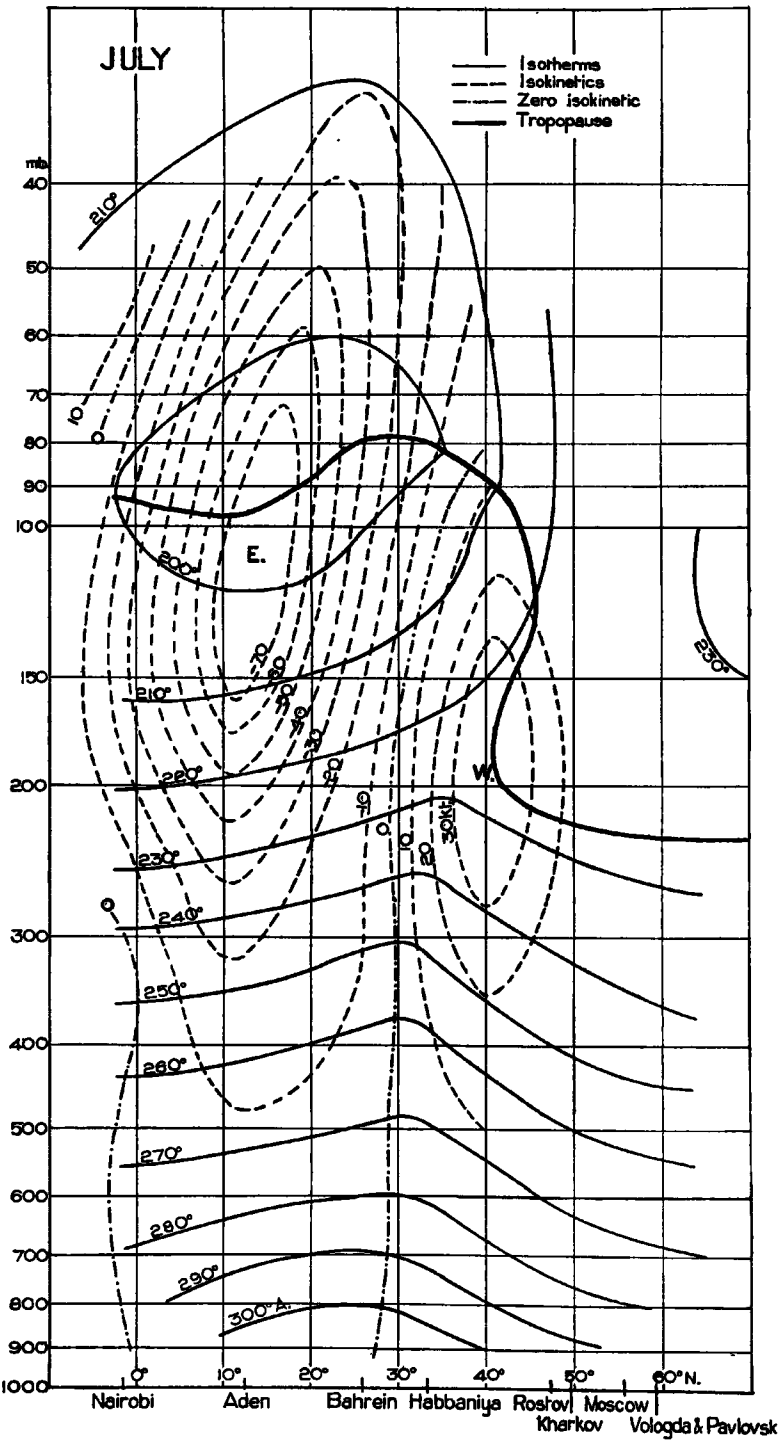


FIG. 4.—ATMOSPHERIC CROSS-SECTION AT 40°E. LONGITUDE SHOWING MEAN TEMPERATURE AND MEAN E.–W. WIND COMPONENT—*continued*

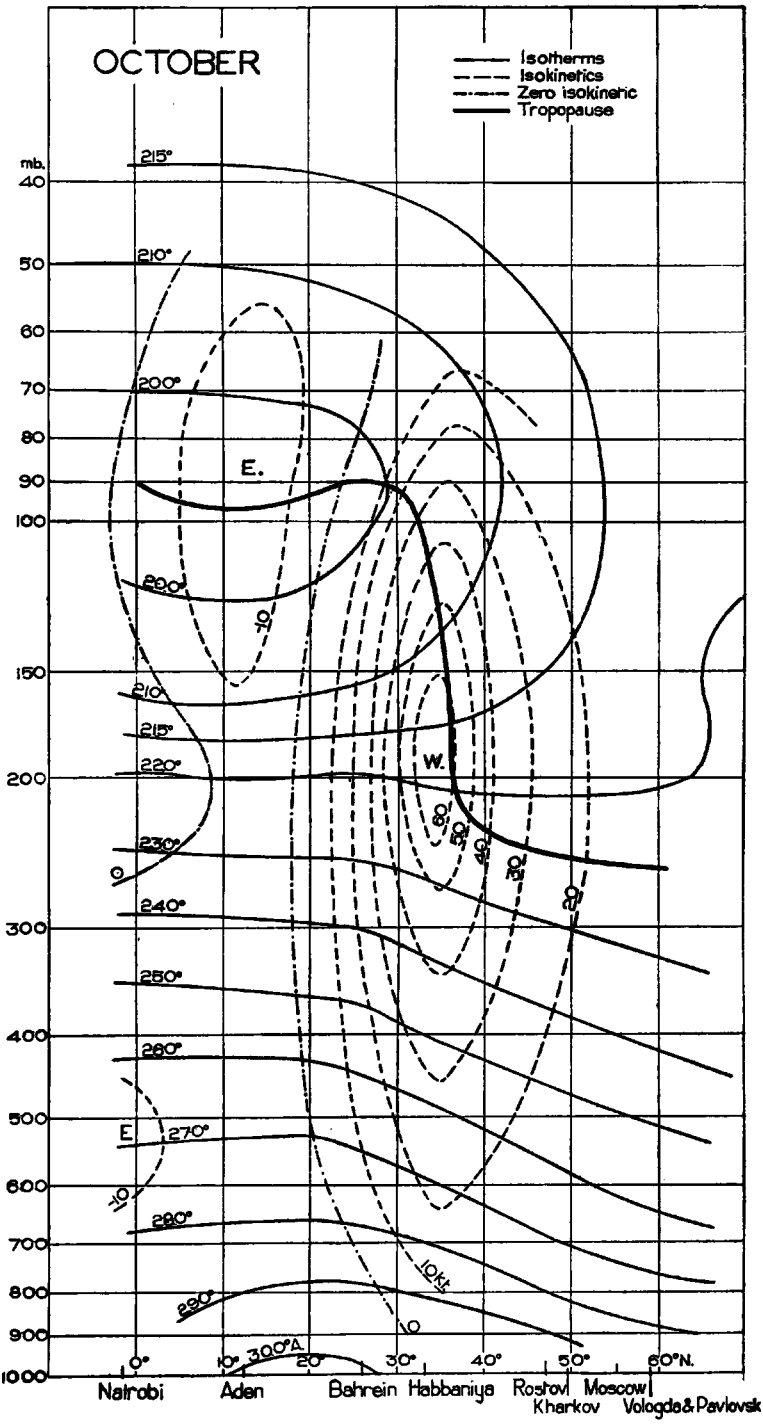


FIG. 4.—ATMOSPHERIC CROSS-SECTION AT 40°E. LONGITUDE SHOWING MEAN TEMPERATURE AND MEAN E.-W. WIND COMPONENT—*continued*

TABLE II—MONTHLY COMPONENTS OF THE WIND SPEED AT STANDARD PRESSURE LEVELS

Habbaniya

Components of the wind from west and south are reckoned positive

Period: 1948–50

	January		February		March		April		May		June		July		August		September		October		November		December	
	W.-E.	S.-N.	W.-E.	S.-N.	W.-E.	S.-N.	W.-E.	S.-N.	W.-E.	S.-N.	W.-E.	S.-N.	W.-E.	S.-N.	W.-E.	S.-N.	W.-E.	S.-N.	W.-E.	S.-N.	W.-E.	S.-N.	W.-E.	S.-N.
mb.	*knots*																							
40	..	..	..	..	..	..	..	..	..	..	..	..	−24·3	+ 3·3	−19·6	+3·0	− 4·5	+ 0·2	..	..	..	..	..	..
60	+31·4	+2·0	..	..	+24·5	+ 1·6	+ 9·8	+ 1·4	+ 0·1	+ 4·8	− 7·1	+ 2·7	−15·1	+ 4·8	−12·7	+4·6	+ 0·6	+ 1·3	+ 6·5	− 0·4	+22·8	−0·3	+25·0	+2·4
100	+45·8	−4·1	+43·4	+4·9	+47·3	+ 3·2	+33·3	+ 4·2	+23·2	+15·3	+20·2	+13·1	+ 1·5	+10·0	+ 5·0	+8·4	+22·7	+12·1	+34·2	+ 7·2	+35·8	+1·8	+40·9	+1·2
150	+65·2	−4·7	+67·2	+4·4	+71·7	+ 5·7	+50·9	+ 7·8	+43·8	+20·4	+41·8	+18·6	+10·0	+ 9·6	+15·8	+9·1	+39·0	+15·4	+57·0	+15·9	+47·7	+4·8	+56·1	+2·2
200	+72·3	−4·0	+77·2	+6·3	+84·3	+12·7	+56·0	+13·1	+49·0	+22·1	+50·7	+19·4	+12·2	+ 7·7	+20·1	+6·9	+43·7	+15·9	+61·6	+18·1	+49·7	+5·6	+67·6	+3·7
300	+62·6	+3·4	+68·1	+7·7	+68·4	+16·3	+46·8	+11·9	+40·2	+18·2	+45·7	+14·0	+ 5·8	+ 3·2	+19·1	+4·3	+40·0	+11·8	+46·1	+11·9	+38·7	+6·5	+58·7	+4·0
500	+33·0	+1·8	+34·5	+6·1	+33·8	+ 8·7	+26·7	+ 7·2	+23·5	+10·0	+22·8	+ 3·2	+ 5·8	+ 5·2	+10·3	+1·8	+18·7	+ 4·2	+23·9	+ 5·7	+17·7	+4·1	+31·1	+3·1
700	+14·0	+1·3	+16·9	+3·3	+17·3	+ 5·3	+13·5	+ 3·2	+16·0	+ 3·7	+14·7	− 3·5	+ 7·4	− 0·6	+ 9·2	−2·4	+15·0	− 1·6	+17·7	+ 2·5	+ 9·5	+3·1	+13·8	+2·5
800	+ 9·0	0·0	+11·0	+2·3	+10·6	+ 2·7	+ 7·6	+ 1·6	+ 8·2	+ 0·2	+11·3	− 8·2	− 9·2	− 7·1	+ 8·6	−6·0	+ 9·4	− 5·8	+10·9	+ 0·7	+ 5·6	+1·8	+ 7·3	+0·4
850	+ 6·8	−0·5	+ 7·8	+1·8	+ 6·8	+ 1·4	+ 4·1	+ 1·3	+ 5·7	− 2·1	+10·3	− 9·4	+10·9	− 9·5	+ 9·3	−7·4	+ 6·8	− 7·1	+ 6·8	− 0·9	+ 3·0	+1·3	+ 4·3	+0·9
900	+ 5·0	−1·2	+ 5·5	+0·9	+ 4·3	+ 0·8	+ 2·4	+ 0·9	+ 4·0	− 3·5	+ 9·2	−11·4	+11·6	−11·7	+ 8·3	−9·3	+ 6·2	− 8·3	+ 3·4	− 2·3	+ 1·3	+0·7	+ 2·3	+0·3
Surface	+ 0·5	−0·3	+ 1·1	+0·4	+ 1·5	+ 0·5	+ 0·1	− 0·2	+ 1·5	− 1·8	+ 3·7	− 2·9	+ 4·0	− 2·2	+ 2·3	−1·4	− 0·6	− 1·0	+ 0·4	0·0	− 0·5	+0·3	0·0	0·0

TABLE III—SEASONAL COMPONENTS OF WIND SPEED AT STANDARD PRESSURE LEVELS

Components of the wind from west and south are reckoned positive

Period: 1948–50

	JANUARY						APRIL						JULY						OCTOBER					
	Nairobi		Aden		Bahrein		Nairobi		Aden		Bahrein		Nairobi		Aden		Bahrein		Nairobi		Aden		Bahrein	
	W.-E.	S.-N.	W.-E.	S.-N.	W.-E.	S.-N.	W.-E.	S.-N.	W.-E.	S.-N.	W.-E.	S.-N.	W.-E.	S.-N.	W.-E.	S.-N.	W.-E.	S.-N.	W.-E.	S.-N.	W.-E.	S.-N.	W.-E.	S.-N.
mb.	*knots*																							
40	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	−40	+3	..	..	..	..	..	..
60	−24	−6	..	..	+26	+12	− 1	0	..	..	+ 5	− 1	+12	+1	−45	0	−40	−1	+ 4	−1	..	..	− 6	−2
100	−11	0	+ 5	0	+57	− 1	− 5	0	+ 4	+2	+38	+ 8	−11	−1	−66	+5	−35	+4	− 3	+1	−17	+6	+ 2	+4
150	−12	−2	+14	−6	+89	−19	+ 1	+1	+18	−1	+66	+17	−31	−3	−71	+4	−23	+2	+ 5	+4	−12	+8	+17	+5
200	−23	+7	+14	−3	+102	−18	− 1	−1	+17	−1	+88	+21	−14	+1	−48	0	−16	+1	+ 6	+3	0	+7	+23	+3
300	−12	−4	+14	−5	+89	−9	+ 8	+1	+16	−1	+59	+12	+ 2	+3	−24	−4	−11	−3	− 3	+4	− 7	+3	+29	+1
500	−13	−1	− 3	−1	+54	0	−13	−1	− 1	0	+28	+5	− 2	−4	− 9	−9	− 9	−4	−12	−1	− 5	−1	+11	−5
700	− 6	−9	− 2	+2	+23	+1	− 7	−3	− 7	−4	+17	+1	− 2	+1	− 1	−3	− 6	−5	− 9	+1	− 8	−12	− 2	−8

Data for Australian stations given by Loewe and Radok[3] and data for Manchuria and Japan published by the Central Meteorological Observatory, Tokyo, have also been utilized. These have been supplemented by observations (extracted from various publications) from aircraft ascents and from sounding balloons, made over many years, for stations in Europe, India and Batavia.

Use of the observational material.—A difficulty which was encountered at the outset in discussing the spatial and seasonal variations of the mean tropopause was the fact that the existing definition of the tropopause is not sufficiently precise for regions in which multiple tropopauses occur. In current meteorological practice, the type and position of the tropopause are specified by the following conventions which are based on those given by W. H. Dines[4] in 1919.

Type I. When the stratosphere commences with an abrupt change in the lapse rate to an inversion the tropopause is at the base of the inversion.

Type II. When the stratosphere commences with an abrupt change to a lapse rate of less than 2°A./Km., or to isothermal conditions, the tropopause is at the level of that change.

Type III. When there is no abrupt change of lapse rate, the tropopause is taken as the level of the point at which the lapse rate becomes less than 2°A./Km., providing that it does not exceed this value for any subsequent kilometre.

It will be noted that with the exception of a Type III tropopause the above definitions imply that the stratosphere is clearly recognizable. In the subtropics where multiple tropopauses are common there is a tendency on the part of some observers to call the inversion at the highest level a Type I tropopause and to neglect any inversions or abrupt changes of lapse rates at lower levels. Other observers on the other hand report as tropopauses levels of abrupt changes which comply with the above definitions, but which are difficult to reconcile with any idea of a boundary surface between troposphere and stratosphere, whilst to complicate the picture other observers exclude all inversions below some arbitrary limit. As a result tropopauses in the subtropics, especially at certain seasons of the year, apparently range haphazardly over pressure levels ranging from 500 mb. to 100 mb. in the course of a few days, and means of such values are very unsatisfactory.

The difficulty of attempting to frame a satisfactory definition can be seen from Fig. 5 which depicts upper air soundings at Habbaniya on February 5, 6 and 8, 1950, after a burst southwards of unusually cold air on the 5th. The tropopauses reported on those days were 130 mb., 500 mb. and 121 mb. respectively.

This drawback in the " mean " level of the tropopause in regions where temperate and tropical tropopauses overlap is well known, and various attempts to overcome this have been proposed. One method which at first sight appears attractive and which has been adopted by many writers, including among others Hess[5], Hutchings[6] and Clarkson[7], is to use the " minimum " tropopause which is less subjective. They make use of the fact that when the mean vertical distribution of temperature above a station is plotted, there is one level (or in rare cases two levels) at which lapse rate is zero. With certain reservations they call this level the " minimum " tropopause, or if there are two such levels they designate the higher as the tropical tropopause and the lower as the polar tropopause. The drawback about this definition is the unfortunate fact that the tropopause as thus defined cannot be reconciled with the average of individual tropopause levels, and, as Hutchings has shown, may exceed the latter by as much as 5 Km.

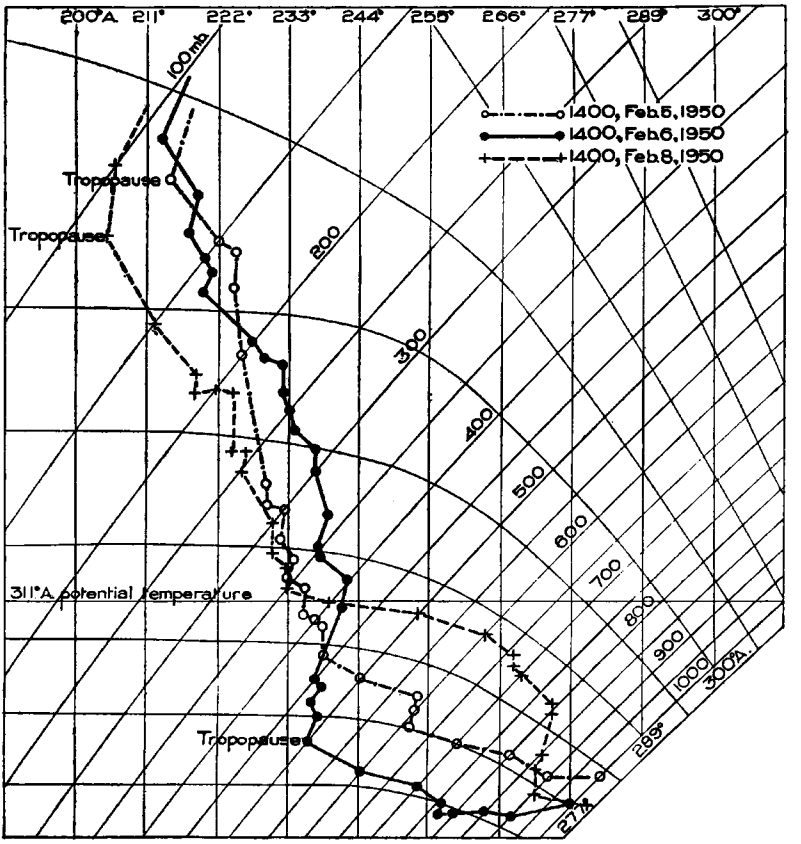


FIG. 5.—REPORTED POSITION OF THE TROPOPAUSE, HABBANIYA, ON FEBRUARY 5, 6 AND 8, 1950, AFTER THE PASSAGE OF A COLD FRONT

For individual soundings this " minimum " definition of the tropopause has little, if any, advantage over the Meteorological Office conventions, for frequently the ascent is ill defined and is easily masked by instrumental errors so that it is impossible to define a strict level of minimum temperature.

J. Bjerknes[8] has stated that at first sight one would place the tropopause simply where the lowest temperature occurs, whilst the true tropopause, which is lower, would be passed over as an insignificant feature in the upper troposphere, and he considers that to arrive at a coherent system it is necessary to admit the diminution in the vertical lapse rate of temperature as a sign of the tropopause.

In a later paper, J. Bjerknes and Palmén[9] regard all sudden decreases of lapse rate in the upper troposphere and lower stratosphere as a sign of the true tropopause, and more recently Flohn and Penndorf[10] have suggested that no attempt should be made to define a single tropopause level, but that a tropopause layer should be substituted extending downwards for a conventional distance from the level at which isothermal conditions are reached.

Despite the imperfections of the " mean " tropopause obtained by the averaging of individual tropopauses, this " mean " tropopause has been used throughout the present paper. In some cases the tropopauses as reported have been rejected, and discontinuities at lower levels selected.

In constructing the meridional cross-sections stations were selected to lie as far as possible along the meridian of 40°E., which has the advantage that all the stations, with the possible exception of Aden, are situated in large continental land masses, and are not subject to the serious difficulties of interpretation which arise when some stations are subject to oceanic and others to continental influences.

Mean east-west and north-south components of the wind at Nairobi, Aden and Bahrein for January, April, July and October are given in Table III, and mean east-west components for Nairobi, Aden, Bahrein and Habbaniya are depicted in Fig. 4 in which the logarithm of the pressure is plotted against the latitude. In this figure lines have been drawn joining all points which have the same speed. Such lines are called isokinetics or isotachs.

Mean temperature at standard pressure levels for the same seasons for Nairobi, Aden, Bahrein, Habbaniya and Rostov are given in Table IV, and mean temperature differences at the levels for the same stations and seasons derived from Table IV are given in Table V. In both tables figures in brackets are based on limited data.

In the present paper actual wind components, in some cases up to 40 mb., have been used for stations south of latitude 35°N., and components of the mean geostrophic wind computed from " Upper winds over the world "[11] have been used north of 35°N.

Mean zonal wind field.—The most significant feature of the mean zonal wind field is the appearance of a pronounced high-level westerly jet stream of over 100 kt. in January at a pressure level of approximately 200 mb. at about 26°N. This jet stream moves very slowly to a position about 29°N. by April, weakens slightly to about 90 kt. and then moves more rapidly to about 40°N. by July when it decreases to about 35 kt. By October it has moved southwards to about 33°N. and has increased in speed to about 65 kt.

A most striking feature of the mean zonal wind field in July is the appearance of an easterly jet stream of over 70 kt. at a pressure level of about 100 mb. at approximately 15°N., with an almost vertical boundary between the easterlies and westerlies at 30°N. This easterly jet stream weakens rapidly by October, and is non-existent in both January and April, its place being taken by westerlies in both the upper troposphere and lower stratosphere.

It will be noted that in drawing the easterly jet stream the centre has been placed at or above the 100-mb. level. The reason for this is that, owing to the strong upper winds at these levels, a large proportion of the balloons failed to reach the levels at 100 mb. and above, and hence the mean components at levels at which the winds were still increasing are too low. An estimate of the true value at 100 mb., from a comparison of the wind speeds at 150 mb. and 100 mb. for all winds which reached at least 100 mb., suggests that the mean wind speed at this level should be increased to about 75 kt. No attempt was made to correct the wind speed at higher levels.

Comparison with cross-sections in the subtropics by other writers shows some important differences, but it is important to note that in the absence of adequate observations of wind at high levels all these writers have based their cross-sections on pressure and temperature data with theoretical computations of winds based on geostrophic assumptions, and the accuracy of such winds near the equator in regions where observations at ground level are scanty, and at upper levels non-existent, is open to question. Hess[5], from mean cross-sections at 80°W. longitude for the combined winter months of January and February and for the combined summer months of July and August for the years 1941 to 1945 inclusive, found in winter a high-level westerly jet stream of about 100 kt.

TABLE IV—MEAN SEASONAL TEMPERATURE

	JANUARY					APRIL				
	Nairobi	Aden	Bahrein	Hab-baniya	Rostov	Nairobi	Aden	Bahrein	Hab-baniya	Rostov
mb.	*degrees Absolute*									
40	208·4	..	(202·3)	210·1	..	207·3	..	207·2	214·6	..
60	197·5	195·8	203·9	210·2	..	198·1	206·1	204·6	212·1	..
100	196·1	194·8	203·7	210·4	(219·0)	196·3	194·8	204·2	212·5	(218·0)
150	208·6	206·6	211·7	215·9	(218·0)	209·4	208·2	213·2	217·2	(218·0)
200	220·9	219·8	219·7	218·6	(218·0)	222·4	221·3	220·6	217·4	(218·0)
300	241·9	241·5	233·2	227·1	222·0	243·7	242·4	236·5	229·8	227·0
400	257·7	256·7	247·4	240·1	..	258·2	257·2	251·1	243·7	..
500	267·1	267·8	258·6	251·2	245·0	267·6	267·8	262·2	255·6	254·0
600	274·7	276·3	267·7	260·5	..	275·7	276·2	271·1	265·1	..
700	283·0	283·4	274·6	267·7	261·0	282·3	284·1	280·2	273·5	269·0
800	286·0	288·1	280·5	273·8	..	288·3	291·6	288·7	281·5	..
850	..	288·8	282·9	276·3	267·0	..	294·2	292·0	285·3	276·0
900	..	290·8	284·8	278·6	..	..	297·1	295·0	289·3	..
Surface	..	298·3	289·8	284·2	..	..	301·1	297·3	294·3	..

TABLE V—MEAN DIFFERENCES OF TEMPERATURE AT STANDARD

	JANUARY				APRIL			
	Nairobi-Aden	Aden-Bahrein	Bahrein-Habbaniya	Habbaniya-Rostov	Nairobi-Aden	Aden-Bahrein	Bahrein-Habbaniya	Habbaniya-Rostov
mb.	*degrees Absolute*							
40	..	..	..	..	..	..	−7·4	..
60	+1·7	−8·1	−6·3	..	−3·5	−3·0	−7·5	..
100	+1·3	−8·9	−6·7	(−8·6)	+1·5	−5·4	−8·3	(−5·5)
150	+2·0	−5·1	−4·2	(−2·1)	+1·2	−5·0	−4·0	(−0·8)
200	+1·1	+0·1	+1·1	(+0·4)	+1·1	+0·7	+3·2	(−0·6)
300	+0·4	+8·3	+6·1	+5·1	+1·3	+5·9	+6·7	+2·8
400	+1·0	+9·3	+7·3	..	+1·0	+6·1	+7·4	..
500	−0·7	+9·2	+7·4	+6·2	−0·2	+5·6	+6·6	+1·6
600	−1·6	+8·6	+7·2	..	−0·5	+5·1	+6·0	..
700	−0·4	+8·8	+6·9	+6·7	−1·8	+3·9	+6·7	+4·5
800	−2·1	+7·6	+6·7	..	−3·3	+2·9	+7·2	..
850	..	+5·9	+6·6	+9·3	..	+2·2	+6·7	+9·3
900	..	+6·0	+6·2	..	..	+2·1	+5·7	..
Surface	..	+8·5	+5·6	..	..	+3·8	+5·0	..

TABLE VI—MEAN MONTHLY TROPOPAUSE

	January		February		March		April		May		June	
	Pres-sure	Tem-perature	Pres-sure	Tem-perature	Pres-sure	Tem-perature	Pres-sure	Tem-perature	Pres-sure	Tem-perature	Pres-sure	Tem-perature
	mb.	°A.	mb.	°A.	mb.	°A.	mb.	°A.	mb.	°A.	mb.	°A.
Nairobi	83·5	191	84·8	192	84·4	191	82·5	191	81·5	191	85·3	194
Aden	90·3	191	90·9	192	85·1	192	85·9	191	87·2	191	92·6	193
Bahrein.. ..	89·7	200	92·0	201	87·9	201	81·7	200	80·6	197	76·7	194
Habbaniya ..	224	216	231	216	235	218	212	216	200	215	88	201
Benina	217	218	238	218	243	219	213	216	197	216	162	213
Nicosia	234	218	253	219	260	219	240	218	218	218	160	215
Malta	235	217	237	216	251	217	224	216	196	214	190	216
Gibraltar ..	221	215	217	215	221	214	210	217	209	219	184	215

AT STANDARD PRESSURE LEVELS

Period: Nairobi, Aden, Bahrein and Habbaniya 1948–50; Rostov 1937–41

JULY					OCTOBER					
Nairobi	Aden	Bahrein	Habbaniya	Rostov	Nairobi	Aden	Bahrein	Habbaniya	Rostov	
degrees Absolute										mb.
211·0	..	(210·2)	209·7	..	213·6	216·1	219·1	215·4	..	40
206·3	204·9	202·0	204·2	..	206·4	205·7	204·7	209·7	..	60
199·7	195·7	199·7	202·3	(222·0)	196·2	194·6	199·2	204·7	(214·0)	100
208·2	207·3	213·6	215·1	(224·0)	207·2	206·0	209·1	210·8	(215·0)	150
219·3	221·1	226·6	228·1	(226·0)	220·6	219·6	220·7	218·7	(218·0)	200
241·1	243·1	248·6	248·3	239·0	242·4	241·1	240·1	235·3	235·0	300
256·9	258·1	262·5	262·5	..	257·5	256·9	254·6	249·6	..	400
266·6	266·7	270·2	271·6	263·0	266·8	267·8	265·9	261·2	258·0	500
274·2	277·4	279·8	278·7	..	274·4	275·0	274·7	270·5	..	600
280·8	288·2	290·2	287·8	278·0	281·5	281·8	282·2	279·9	275·0	700
284·6	296·5	298·8	296·3	..	286·3	289·9	290·8	288·0	..	800
..	300·1	303·1	300·5	290·0	..	293·4	294·6	291·7	278·0	850
..	303·2	306·8	304·8	..	..	296·4	298·2	295·2	..	900
..	305·9	306·9	306·9	..	..	301·3	301·1	296·4	..	Surface

PRESSURE LEVELS CONSTRUCTED FROM TABLE IV

Period: Nairobi, Aden, Bahrein, Habbaniya 1948–50. Rostov 1937–41

JULY				OCTOBER				
Nairobi-Aden	Aden-Bahrein	Bahrein-Habbaniya	Habbaniya-Rostov	Nairobi-Aden	Aden-Bahrein	Bahrein-Habbaniya	Habbaniya-Rostov	
degrees Absolute								mb.
..	..	+0·5	..	−2·5	−3·0	+3·7	..	40
+1·4	+2·9	−2·2	..	+0·7	+1·0	−5·0	..	60
+4·0	−4·0	−2·6	(−19·7)	+1·6	−4·6	−5·5	(−9·3)	100
+0·9	−6·3	−1·5	(−8·9)	+1·2	−3·1	−1·7	(−4·2)	150
−1·8	−5·5	−1·5	(+2·1)	+1·0	−1·1	+2·0	(+0·7)	200
−2·0	−5·5	+0·3	+9·3	+1·3	+1·0	+4·8	+0·3	300
−1·2	−4·4	+0·0	..	+0·6	+2·3	+5·0	..	400
−0·1	−3·5	−1·4	+8·6	−1·0	+1·9	+4·7	+3·2	500
−3·2	−2·4	+1·1	..	−0·6	+0·3	+4·2	..	600
−7·4	−2·0	+2·4	+9·8	−0·3	−0·4	+2·3	+4·9	700
−11·9	−2·3	+2·5	..	−3·6	−0·9	+2·8	..	800
..	−3·0	+2·6	+10·5	..	−1·2	+2·9	+13·7	850
..	−3·6	+2·0	..	..	−1·8	+3·0	..	900
..	−1·0	+0·0	..	..	+0·2	+4·7	..	Surface

PRESSURE AND TEMPERATURE

Period: 1948–1950

July		August		September		October		November		December			
Pressure	Temperature	Pressure	Temperature	Pressure	Temperature	Pressure	Temperature	Pressure	Temperature	Pressure	Temperature		
mb.	°A.	mb.	°A.	mb.	°A.	mb.	°A.	mb.	°A.	mb.	°A.		
95·0	195	94·0	197	93·8	196	91·7	193	83·8	191	78·0	191	..	.. Nairobi
96·3	194	96·2	195	96·5	194	97·8	193	92·9	192	92·5	192	..	.. Aden
78·2	194	85·5	194	84·6	194	90·0	196	91·0	200	94·0	201	..	.. Bahrein
83	198	90	199	97	199	150	205	184	210	202	211	..	Habbaniya
93	201	97	202	107	206	151	210	163	210	206	214	..	.. Benina
97	203	98	203	99	204	125	209	164	211	221	215	..	.. Nicosia
123	210	119	210	142	211	177	212	200	212	228	216	..	.. Malta
137	211	137	211	169	212	189	213	174	209	224	217	..	Gibraltar

(50 m./sec.) at about latitude 35°N. and at a height of about 12 Km., whilst in summer it occurred at latitude 55°N. and its speed had decreased by a half. Chaudhury[12], who prepared a cross-section for January and February over India and Pakistan along the meridian of 76°E. from one year's data only, found three westerly jet streams, one between 15°N. and 20°N., another to the south of the Himalayas at about 31°N., and another to the north of the Himalayas.

Cross-sections prepared by Loewe and Radok[3] for the south-west Pacific in longitude 150°E., a mainly continental régime, show a winter jet stream of about 105 kt. (120 m.p.h.) at about 25°S. at a height of about 12 Km., and a summer jet stream about 45 kt. (50 m.p.h.) at 40°S. at a height of about 10 Km. The winds used in their cross-sections were components of the mean geostrophic wind, and a comparison which they carried out between such computed winds and winds observed by radar at three stations between latitudes 19°S. and 34°S. in Australia, showed that the observed winds were only 55 per cent. of the computed winds. In a later note Loewe and Radok[13], from new Tasmanian data, suggest that the summer maximum should be twice as high as that given in their earlier paper.

Hutchings[6], who prepared a meridional cross-section for a purely oceanic region in the south-west Pacific at 170°E., found from the computed geostrophic wind components a single maximum of 100 kt. at 30°S. in winter, and in summer a maximum of 60 kt. at 46°S. and a secondary maximum of 50 kt. at 31°S.

Mean zonal temperature field and tropopause.—The most important features of the mean zonal temperature field are the concentration of latitudinal temperature gradients, the baroclinic zone in the troposphere beneath the westerly jet stream at all seasons of the year, and in July a second baroclinic zone south of 30°N. It can be seen that the latitudinal gradients of temperature vanish at a pressure level of about 200 mb., whilst above this level the gradients of temperature are reversed, and temperature in the lower stratosphere above the cold air to the north is warmer than temperature in either the lower stratosphere or upper troposphere above the warm tropical air. The lowest temperature occurs in the lower stratosphere above the thermal equator, which in January is at or near the geographical equator, and in July is at approximately 30°N.

The tropopause over the " warm " regions occurs between pressure levels of 100 mb. and 80 mb. whilst over the " cold " regions to the north the tropopause occurs between pressure levels of 250 mb. and 200 mb. ; Table VI shows the monthly values of the mean tropopause pressure and temperature. It can be seen that at Nairobi the tropopause is highest in the winter and spring at a pressure level of about 80 mb., and lowest in the summer and autumn at a pressure level of about 90 mb. At Aden the tropopause is highest in March and April at a pressure level of about 85 mb., and lowest in September and October at a pressure level of about 95 mb. At Bahrein the tropopause is at its highest in June at a pressure level of less than 80 mb., and lowest in December at a pressure level of about 95 mb., whilst at Habbaniya the tropopause occurs at a pressure level of 83 mb. in July and of about 220 mb. in January.

These figures show that as the sun moves northward there is a close relationship between the height of the tropopause and the position of the thermal equator, and that the highest tropopause occurs at the thermal equator, which is not unexpected. What is unexpected, however, is the fact that as the thermal equator moves south the height of the tropopause apparently falls with it. This, if real, suggests that the simple pattern of meridional circulation, in which air ascends at the thermal equator south of the jet stream and subsides north of the jet stream, requires modification when the sun is moving southwards. In support of this suggestion, the differences between the north-south

components of the wind at Nairobi and Aden from Table III show that in April there is little or no horizontal divergence or convergence at Aden at levels between 700 mb. and 100 mb., whereas in October there is marked horizontal divergence at the 200-mb., 150-mb. and 100-mb. levels.

The nature of the transition between the high-level tropopause and the low-level tropopause is also obscure, and it is uncertain whether in the baroclinic zone the tropopause should be regarded as a continuous surface or as a series of discontinuous surfaces. Ramanathan and Ramakrishnan[14], who first recognized the discontinuous nature of the tropopause surface over northern India in 1933, attributed this to the large-scale folding of the tropopause resulting from meridional circulations in the vertical plane. A study of the day-to-day soundings at Habbaniya shows that in January there is one tropopause at about 220 mb., and on about one day in two there is also a second tropopause at about 90 mb. In February and March conditions are somewhat similar, although in February the proportion of upper tropopauses appears to be rather less. In April and May there appear to be at least three tropopauses at pressure levels of approximately 220 mb., 160 mb. and 80 mb., whilst on many occasions there are up to five. In June, July, August, September, and most of October, observations show that there is only one tropopause at a pressure level between 100 mb. and 80 mb. In late October and November the same multiplicity of tropopauses as in April and May occurs, whilst in December there appear to be two tropopauses with the lower predominating at about 220 mb. At Bahrein observations show that there is always a high tropopause between 100 mb. and 80 mb., but in December and January there is also evidence of the low temperate tropopause. In January therefore the tropical tropopause at 100–80 mb. and the temperate tropopause at 220–200 mb. overlap by at least 500 miles.

Observations from stations in the Mediterranean show that there is a marked longitudinal variation of the tropopause. At Benina and Nicosia the variation of the tropopause is similar to that at Habbaniya, although the transition from temperate to tropical tropopause occurs slightly later and is less abrupt than at Habbaniya. At Gibraltar the temperate tropopause persists throughout the whole year, and from February to mid June there is little evidence of an upper tropopause. From July to January there appear to be two tropopauses, but in July and August the tropical tropopause is the most marked. The observations from Malta are intermediate between those from Gibraltar and Nicosia. These general conclusions may be compared with those of Nagai and Matuoka[15] who investigated the tropopause from radio-sonde ascents at Tateno (36°03′N., 140°08′E.) from January 1948 to April 1949, and found that in the winter and early spring double tropopauses occurred at heights of about 10 Km. and 15½ Km., whilst from March to November there was only a single tropopause which rose rapidly from under 12 Km. in April to over 14 Km. in June, and then more gradually to over 16 Km. in September, October and November—the highest value occurring in September.

Fig. 6, which depicts the mean monthly temperature at Habbaniya at standard pressure levels and the mean monthly tropopause pressure and temperature during five of the first seven months of the year, shows that during the period from January to May the temperature increase, due to the surface heating, is distributed evenly throughout the troposphere between the surface and about 200 mb., whilst in the stratosphere above this level the thermal structure is very stable and shows very little change. Between May and June, however, the continued rise in temperature in the troposphere is accompanied by a marked cooling in the lower stratosphere at about 100 mb., so that the low tropopause, which originally formed a complete barrier to penetrative convection from below, is easily broken through with the result that the surface heating

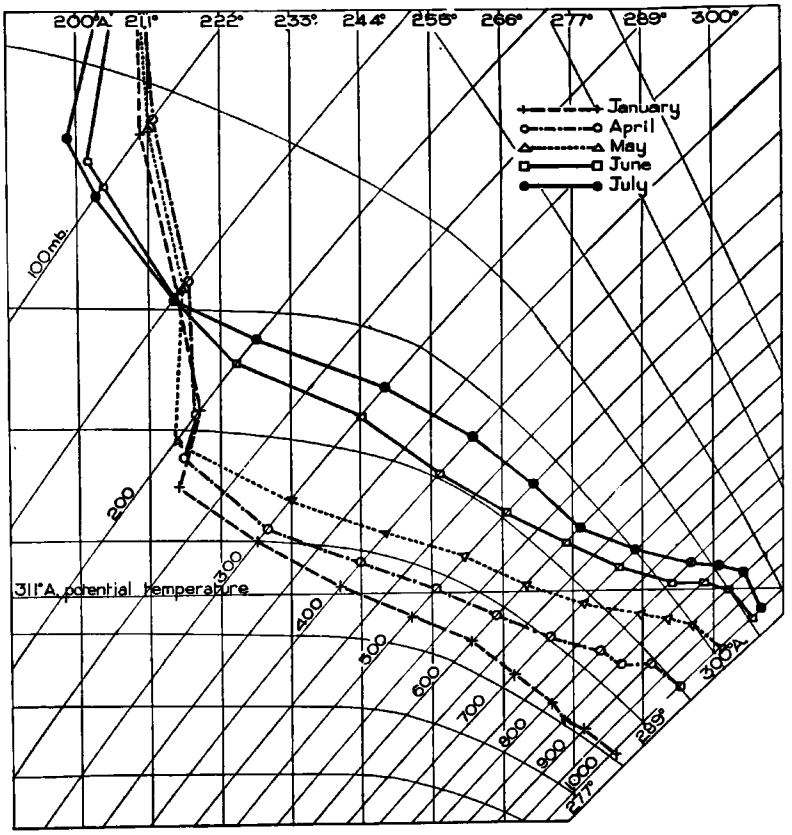


FIG. 6.—MEAN TEMPERATURE, HABBANIYA

is distributed through a deeper layer of the atmosphere, thereby producing the discontinuity in the rise of the maximum temperature shown in Fig. 2. H. Arctowski[16], who first noted temperature discontinuities of this kind at Melbourne and elsewhere, described them by stating that the natural course of the temperature associated with local conditions appeared to be interrupted, just as if the station had been lifted into another climate similar to but colder than it might be during the summer. An analysis of the breakdown in May 1949, with the aid of Fig. 7, shows that from the 1st to the 23rd the temperature increased at all pressure levels from the surface to approximately 200 mb., whilst above 200 mb. there was little change apart from a slight cooling of perhaps 7°F. at 100 mb. Between the 23rd and 24th however there was very rapid cooling in the lower stratosphere, the temperature at the 90-mb. level now being some 20°F. lower than on the 1st. By the 25th this cooling had extended downwards to the 150-mb. level, accompanied by very rapid warming between 260 mb. and 200 mb. which appeared to be due to subsidence. The warming of the lower layers continued between the 25th and 26th, and by the latter date the inversion at 260 mb. appeared only as an insignificant feature of the tropopause, which even if real no longer constituted any barrier to penetrative convection from below, leaving only an upper tropopause at the level of maximum cooling.

It has been noted by several writers, e.g. Bjerknes and Palmén[17] and Kraus[18], that in Europe folded tropopauses occur with all major warm frontal surfaces,

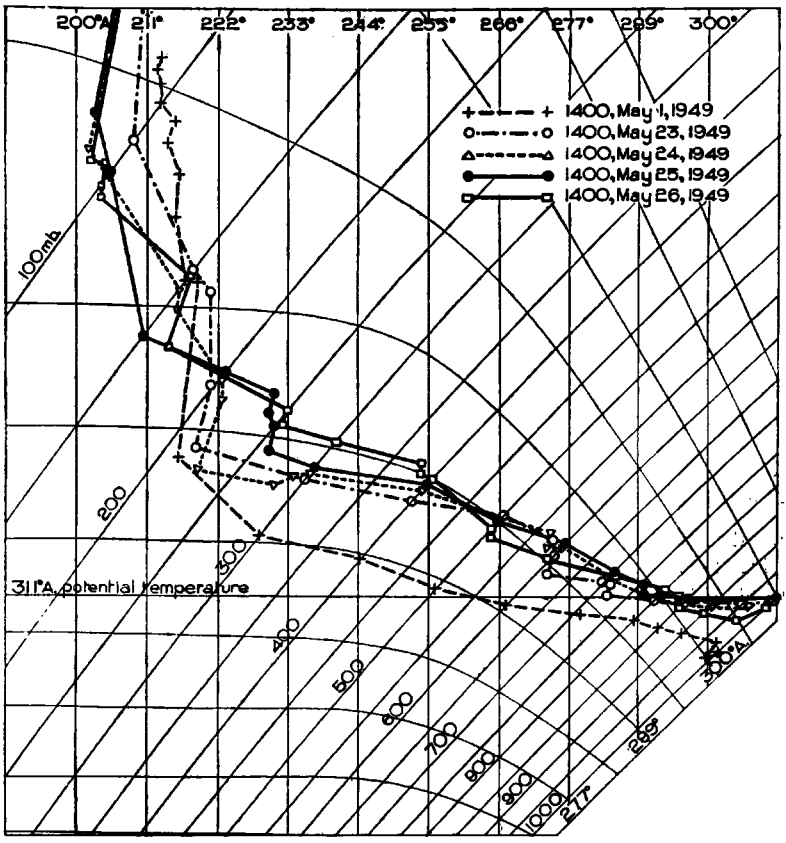


FIG. 7.—BREAK-DOWN OF LOWER TROPOPAUSE AND FORMATION OF UPPER TROPOPAUSE, HABBANIYA, MAY 1949

and that the first formation of the folded tropopause appears to take place about 500–1,000 miles ahead of the surface warm front. As the thermal equator moves north, the baroclinic zone is a warm-front transitional zone, and hence it might be anticipated that a similar characteristic folding of the tropopause would occur. On the other hand instead of cooling by advection it is possible that the cooling at high levels might be the result of horizontal divergence as suggested by Goldie[19] in which case the tropopause surfaces would be discontinuous.

It does not appear to be possible either from the observations at Habbaniya or the rather scanty observational network in this area to decide between these alternatives. Although of theoretical importance it does not appear to be of very great practical importance which theory is correct, and in view of the very large number of multiple tropopauses encountered at Habbaniya in the transitional months, it was considered best, at least on aesthetic grounds, to depict the tropopause on the seasonal cross-sections as a continuous folded surface.

The spatial and seasonal variations of the mean tropopause, together with the position of the jet stream taken from " Upper winds over the world "[11], are shown in Fig. 3. These charts which are based upon all the data available are believed to be reasonably reliable and indicate that the main features of the meridional cross-section at 40°E. apply with minor modifications to the whole

of the African and Asian continents. No attempt was made on these charts to depict either a folded or discontinuous tropopause, and the tropopause is shown as a steeply sloping continuous surface to the north of the jet stream.

Jet stream and latitudinal gradient.—There is at present no generally accepted theory of the formation of the jet stream. Rossby and collaborators[20] argue that the transfer of absolute vorticity by large-scale lateral mixing could account for some of the more important features of the jet stream, and that the latitudinal variation of temperature is dynamically produced by the kinetic energy of the jet stream. Palmén and Newton[21] consider that the jet stream is always closely associated with the polar front. Namias and Clapp[22] have advanced their confluence theory (kinematic frontogenesis), according to which the jet stream is caused by the gradual flowing together of a current of warm mid-tropospheric air curved anticyclonically from the south and of cold mid-tropospheric air curving cyclonically from the north. Bleeker[23], however, whilst recognizing the importance of confluence, considers that physical frontogenesis along a line of differential heating is the main cause of the jet stream.

Of these theories, Bleeker's appears to afford the most likely explanation of the African-Indian high-level jet stream.

The intimate relationship between the mean jet stream and the integrated mean latitudinal temperature gradient can be readily demonstrated. In the case of steady geostrophic flow

$$2\omega u \sin\phi = -\frac{1}{\rho}\frac{\partial p}{\partial y} \qquad \ldots\ (1)$$

$$g = -\frac{1}{\rho}\frac{\partial p}{\partial z} \qquad \ldots\ (2)$$

$$p = R\rho T, \qquad \ldots\ (3)$$

where u is the velocity along the west-to-east axis Ox, Oy is the south-to-north axis, z is measured vertically upwards, p, ρ and T are the pressure, density and temperature respectively of the air, R is the gas constant and $2\ \omega \sin\phi$ is the Coriolis coefficient. From (2) and (3)

$$\frac{g}{RT^2}\frac{\partial T}{\partial y} = \frac{\partial}{\partial y}\left(\frac{1}{p}\frac{\partial p}{\partial z}\right) = \frac{\partial}{\partial z}\left(\frac{1}{p}\frac{\partial p}{\partial y}\right) \qquad \ldots\ (4)$$

From which if $(\partial T/\partial y)_p$ is the latitudinal variation of temperature along a pressure surface

$$\frac{g}{RT^2}\left\{\left(\frac{\partial T}{\partial y}\right)_p + \frac{\partial T}{\partial p}\frac{\partial p}{\partial y}\right\} = -g\rho\frac{\partial}{\partial p}\left(\frac{1}{R\rho T}\frac{\partial p}{\partial y}\right)$$

$$= -\frac{g\rho}{RT}\frac{\partial}{\partial p}\left(\frac{1}{\rho}\frac{\partial p}{\partial y}\right) + \frac{g}{RT^2}\frac{\partial p}{\partial y}\frac{\partial T}{\partial p}.$$

$$\text{whence } R\left(\frac{\partial T}{\partial y}\right)_p = -p\frac{\partial}{\partial p}\left(\frac{1}{\rho}\frac{\partial p}{\partial y}\right), \qquad \ldots\ (5)$$

and hence from (1) and (5)

$$2\ \omega\ u \sin\phi = R\int_p^{p_0}\left(\frac{\partial T}{\partial y}\right)_p \frac{dp}{p} \qquad \ldots\ (6)$$

and hence the maximum wind occurs at the pressure level at which $(\partial T/\partial y)_p$

= O (i.e. the level at which the northward gradient of temperature along an isobaric surface is zero) which level, from the mean meridional cross-sections, is about 200 mb.

The winds at 200 mb. calculated from equation (6) for latitudes 26°N. and 40°N., using the temperature differences between Bahrein and Habbaniya for January and between Habbaniya and Rostov for July respectively from Table V, are 105 kt. and 40 kt. This very close agreement is probably accidental, but the general agreement suggests that the thermal gradient could be the cause of the jet stream, whilst the fact that the jet stream appears on the cross-sections from January to July as a weakening core of fast-moving air with a large shear on both sides suggests that it is subject to decay by the action of large-scale horizontal eddies which mix the fast-moving thermally induced stream with stationary air to the north and south.

Relation with the onset of the Indian monsoon.—As early as 1686, Halley[24] stated that the cause of the monsoon was due to differential heating, but until recently no very satisfactory explanation had been given for the retardation of its onset over India as compared with Burma, and why it subsequently advances over India in a spectacular manner.

Maung Tun Yin[25], who carried out a study of the onset of the monsoon, suggested that its retardation over India as compared with Burma could be explained by the fact that during the winter and spring months the seasonal flow patterns at high levels showed that there was an upper trough line, presumably orographically determined, orientated north-south at about 85°E., which would accelerate the monsoon over Burma which lies to the east of this upper trough, whilst retarding it over India which lies to the west of the trough. The mean-flow patterns at high levels over India show that the upper trough in summer is located at 75°E. Maung Tun Yin found that the displacement of this upper trough, which occurred relatively abruptly, appeared to be related to the advance of the monsoon over India, and pointed out that the displacement of this trough westward could explain the advance of the monsoon, or alternatively that the advance of the monsoon could explain the displacement of the trough westwards. As the latter precluded an explanation of the onset of the monsoon, he investigated the possibility of explaining the shift of the upper trough westwards by some other mechanism. Using values of the geostrophic wind at 500 mb. for five-day intervals averaged over the whole of Eurasia from 0° to 130°E., he found that the region of strongest westerlies, about 24 kt. at the beginning of May, was situated at about 30°N., and that this region moved to about 40°N. in the last week of May and disintegrated rapidly during the first week of June. From this he argued that the westerly stream, which in the pre-monsoon season had circled the Himalayas along its southern border, was displaced to the north of the Himalayas during this period, and that this resulted in a rapid orographically determined displacement westwards of the upper trough.

Whilst this explanation is undoubtedly ingenious there is little evidence to show that the upper-level trough in summer is orographically determined, and his deductions, from geostrophic winds averaged from 0° to 130°E. over a region which embraces the Mediterranean, the Iraq and Persian deserts, the Himalayan plateau and Indo-China where pressure régimes vary considerably and where, especially in the area in question, observations are scanty, are not altogether convincing.

According to the Habbaniya data, the westerly component of the wind at the 200-mb. level falls rapidly from a maximum of 84 kt. in March to 56 kt. in April, then more slowly to 49 kt. in May, increases to 51 kt. in June and falls off rapidly to 12 kt. in July. This pattern is repeated at the 300-mb. level, whilst at the 500-mb. level there is a slow decrease from 34 kt. in March to 27 kt. in April, to 23½ kt. in May, to 23 kt. in June, and then a rapid fall to 6 kt. in July, and this pattern shows little variation from year to year.

As the jet stream is related to the latitudinal temperature gradient, any weakening in the thermal field should be reflected in the strength of the jet stream. Since the jet stream must be near but still south of Habbaniya when the abrupt change of tropopause occurs, there should be a slight intensification of the temperature gradient southwards and hence a slight increase in the strength of the jet stream. It would thus appear from the Habbaniya data that the jet stream moves through Habbaniya late in June and that the collapse of the jet stream does not take place until some time after the " burst of the monsoon " which usually occurs at the end of May.

In a discussion of the tropopause observations in May 1949 it was shown that the change in tropopause level occurred on May 27. In 1948 it occurred on June 9, and in 1950 on May 26. The date of change in 1950 given by Dewar[1] from the reported tropopauses was May 19, but an examination of the original data shows that, although the upper tropopause was well developed by this date, the lower tropopause still constituted an effective barrier to a break-through from below until May 26.

The dates of the " burst of the monsoon " along the Malabar coast during these years, extracted from the *Indian Daily Weather Reports*, are as follows :—

1948, June 10 ; 1949, May 23 ; 1950, May 27.

It is suggested therefore that the sudden onset of the monsoon over India results from the break-down of the lower tropopause barrier due to the combination of the rapid heating of the layers between the surface and 200 mb. and the cooling, either by advection or dynamical effects, of the layers above, and that the onset of the monsoon precedes the movement of the jet stream to a position north of the Himalayas.

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